Global Tectonic Setting and Climate of the Late Neoproterozoic: A Climate-Geochemical Coupled Study

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Whereas the snowball Earth hypothesis seems to account for most of the major features of the Neoproterozoic glacial records, the causes that drove the Earth into a snowball state remain largely open to debate. Most of the mechanisms leading to the initiation of a snowball Earth are based on the existence of the unusual preponderance of land masses in the tropics. However, the time of the youngest Neoproterozoic glaciation is characterised by a rather widely distributed geography from low-to-high latitudes. In the absence of reliable knowledge of Neoproterozoic topography, two series of coupled ocean-atmosphere climate model simulations were carried out with a Late Neoproterozoic paleogeography (580 Ma) and solar luminosity reduced by 6% relative to today, the first one with flat continents and the second one with mountain ranges that mimic the Pan-African Orogen occurring at this time. Those climatic simulations coupled to the long-term carbon cycle have allowed to better constrain the atmospheric pCO₂ and the associated climate by the time of the youngest late Proterozoic glaciation. The Pan-African Orogen runs result in a snow accumulation pattern compatible with a regional-scale glaciation more less extensive while the no relief runs do not succeed in initiating any glaciation. These results could give additional support to the inferences from many authors that some of the glacial deposits originally attributed to a snowball-like glaciation could in fact be the consequence of a more localised glaciation due to the important orogen occurring at the end of the Neoproterozoic.

1. INTRODUCTION

The Earth underwent at least two episodes of severe glaciation at the termination of the Proterozoic era, one around 730 Ma (Sturtian/Rapitan episode), and a second one around 600 Ma (Marinoan/Varangian episode). In 1992, Kirschvink [1992] suggested that both episodes might correspond to a complete glaciation event (the “snowball Earth” hypothesis), with total
cover-up of continental and oceanic surfaces by ice. This hypothesis seems to account for most field observations, and especially for the low paleolatitudes glacial deposits inferred from paleomagnetic results (see [Donnadieu et al., 2003] and [Pollard and Kasting, 2003]) as well as for the depletion in 13C observed in the cap carbonates overlying the glacial deposits [Donnadieu et al., 2003; Hoffman et al., 1998a; Hoffman et al., 1998b]. Hoffman and colleagues interpret this negative shift in δ13C to be the consequence of prolonged low organic productivity during snowball events, high rates of carbonate sedimentation due to the high alkalinity and carbon fluxes into the ocean as a result of enhanced weathering during the “super-greenhouse” climate in the aftermath of the snowball glaciation. Whereas the snowball Earth hypothesis seems to account for most of the major features of the Neoproterozoic glacial records, the causes that drove the Earth into a snowball state remain largely open to debate. Most of the mechanisms leading to the initiation of a snowball Earth are based on the existence of the unusual preponderance of land masses in the tropics (see below the part 2). However, if a general consensus has merged favouring a low to mid latitude supercontinent named ‘Rodinia’ for the 1100–800 Ma period [Dalziel, 1997; Meert, 2001; Weil et al., 1998] that would have broken up after 800 Ma [Meert and Torsvik, 2003; Torsvik et al., 2001], there are no reliable paleogeographic reconstructions for the 700–600 Ma time period [Meert and Powell, 2001] because the paleomagnetic data are poorly constrained or entirely lacking [Evans, 2000]. Nevertheless, the few available recent geochronologic studies seem to indicate that the age of the last glaciation of the Neoproterozoic is most reasonably equals to 600 Ma [Barfod et al., 2002; Gorokhov et al., 2001; Thompson and Bowring, 2000]. Consequently, many climate modellers have studied the younger snowball event adopting a paleogeographic reconstruction typical of the 580 Ma time period [Chandler and Sohl, 2000; Donnadieu et al., 2002] or even of the 545 Ma time period [Crowley et al., 2001; Hyde et al., 2000]. Those reconstructions, based on Laurentia migration to high latitudes by 577 Ma [Torsvik et al., 1996], show the paleolatitudes of the vast majority of the continental masses range from the pole to the equator. [Hoffman and Schrag, 2002] suggest that those continental configurations might have ended the snowball era rather than have favour it. It seems however interesting to test and to quantify the effect of such a paleogeography on the Neoproterozoic climate via its impact on the climate-silicate weathering feedback. Indeed, in the absence of any proxy, the critical climatic parameter, i.e. the atmospheric pCO₂, has been defined as an arbitrary boundary condition, most previous climatic and geochemical studies assuming a low pCO₂ prior to the glaciation, corresponding to an already severe cooling. However, the atmospheric CO₂ level is completely defined by the balance between the volcanic degassing and the continental silicate rock weathering [Goddéris and François, 1995; Walker et al., 1981]. Although the volcanic degassing rate is poorly constrained for Neoproterozoic times, we will explore the impact of the continental configuration and of the presence of mountainous area (the Pan-African orogeny) on the continental weathering rates, hence on background pCO₂ and climatic setting for Marinoan times prior to any snowball event. In this contribution, we have performed two suites of global climate model experiments with a solar constant reduced by 6% relative to present day value and with various fixed values of CO₂, the first set of runs is for the 580 Ma continental reconstruction with no relief (called after NR) and the second set of runs is for the same paleogeography but with two mountain ranges resulting from the Pan–African Orogen (called after PAO). Each experiment is run for thousand years to climatic equilibrium. In this way, we have evaluated the sensitivity of the critical value of atmospheric CO₂ leading to a full glaciation to the existence of mountain ranges. Then, we have used those climatic simulations to run the global geochemical model COMBINE to calculate the steady state atmospheric CO₂ level. This modelling approach has been performed in order to answer the two following key questions: (1) Can we constrain the atmospheric pCO₂ by the time of the Marinoan glaciation? (2) Once the pCO₂ level is constrained, how does the corresponding global climate looks, and particularly what kind of glaciation can be expected (global or rather regional)?

2. SHORT REVIEW OF THE PLAUSIBLE TRIGGERS

Kirschvink [1992] was the first to postulate that conditions amenable to global glaciations were set up by an unusual preponderance of land masses within middle to low latitudes as it would lead to lower tropical temperatures, through reduced receipt of shortwave radiation and a smaller tropical greenhouse effect. In addition, placing more continents in the tropics may induce an increase of the silicate weathering rate, leading to an increase of atmospheric CO₂ consumption. Such an equatorial continental configuration turns out to be effective only for Sturtian times. In two recent studies, following these lines of evidence, Goddéris et al. [2003] and Donnadieu et al. [2004a] have quantified the possible connections between the tectonic forcing and the ice house state of the Neoproterozoic. The older snowball glaciation (730 Ma, the Sturtian one) is contemporary with the dislocation of the Rodinia supercontinent which is heralded and accompanied by intense magmatic events, including the onset of large basaltic provinces spanning the 825–755 Ma time interval ([Goddéris et al., 2003], see references therein). Then, because basaltic rocks weather about eight times faster than granitic rocks [Dessert et al., 2003], the global weatherability of the continental sur-
face drastically increases following the set up of basaltic traps, resulting in an enhanced consumption of atmospheric CO₂, inducing a global long term climatic cooling. Based on the best estimates of the size and of the location of the Laurentian magmatic province, [Goddéris et al., 2003] show that more than 140 ppm of atmospheric CO₂ could be consumed via this mechanism which is sufficient to trigger a snowball glaciation, assuming a pre-perturbation pCO₂ value of 280 ppm. On the other hand, exploring the relationship between the paleogeographic evolution and the atmospheric CO₂ level through the coupling of the climate model CLIMBER-2 and the geochemical model COMBINE, Donnadieu et al. [2004a] demonstrate that the dislocation of the supercontinent ‘Rodinia’ could have promoted a CO₂ decrease of more than 1000 ppm. This is the result of the increase in continental runoff (following the dislocation) which results in an enhanced consumption of atmospheric CO₂ by the silicate weathering and subsequent trapping of carbon within carbonate sediments. Coupled to the enhanced weatherability due to the onset of large magmatic provinces, the dislocation of the Rodinia supercontinent might have driven the Earth into a snowball glaciation.

The hypothesis proposed by Schrag et al. [2002] also relies on the equatorial location of continental masses to explain the Marinoan snowball glaciation, although the continental configuration appears to be more dispersed in latitude by the time of the younger global glaciation. This hypothesis invokes methane release from organically-rich sediments accumulated within warm equatorial river deltas into a low oxygen atmosphere, ensuring a rather long residence time of CH₄ within the atmosphere, boosting consumption of atmospheric CO₂ by weathering of continental silicates through enhanced greenhouse conditions. Such methane release might explain the pre-glaciation long term decrease in seawater δ¹³C documented prior to the Marinoan glaciation. Once methane release ends up, atmospheric methane collapses through oxidation and the remaining atmospheric CO₂ might be low enough to result in the onset of a snowball glaciation. Following this logic but using a one-dimensional photochemical model, [Pavlov et al., 2003] re-explore the chemistry of the Proterozoic atmosphere. They suggest that Proterozoic O₂ levels were significantly less than today’s concentration based on recent studies using sedimentary sulfur isotope ratios [Canfield, 1998; Canfield and Teske, 1996] and trace sulfates in carbonates [Hurtgen et al., 2002]. Then, arguing that the methane production from the oxygen-poor Proterozoic ocean could have been 10–20 times higher than the modern one, they show that it is sufficient to maintain 100–300 ppm of methane in the atmosphere. The key question to be answered is the nature of the mechanism that ends the methane release, leading to the snowball glaciation.

3 THE MODELS AND THE EXPERIMENTS

3.1. The Climate Model and the Experiments

The coupled ocean-atmosphere model CLIMBER-2 is used extensively to investigate present, future and past climates. The CLIMBER-2 model has been fully described in [Petoukhov et al., 2000] and successfully simulates the last glacial/interglacial cycle [Ganopolski et al., 1998b; Ganopolski and Rahmstorf, 2001; Ganopolski et al., 1998a] as well as earlier climates [Donnadieu et al., 2004b]. Briefly, the atmospheric module is a 2.5-dimensional statistical-dynamical model and has a resolution of 10° in latitude and approximately 51° in longitude (7x18 grid). The ocean module is based on the (averaged) equations of Stocker et al. [1992] and describes the zonally averaged characteristics for the ocean realm with a latitudinal resolution of 2.5°. The model also includes a thermodynamic sea-ice model. More details on the changes and on the limits of the CLIMBER-2 model when simulating very different climates, such as the Neoproterozoic one, than the present-day one can be found in Donnadieu et al. [2004b].

For the climatic simulations, shared boundary conditions are: (1) A solar luminosity 6% below present. (2) The Earth’s orbit about the sun is circular (eccentricity = 0) and the Earth’s obliquity is 23.5°. This setting causes an equal receipt of solar insolation for both hemispheres. (3) The land-sea distribution is the 580 Ma paleogeography used and described in Donnadieu et al. [2002]. (4) In addition, because river drainage basins are unknown for this time period, we defined a river mask in which rain falling and snow melt on land is equally redistributed to all coastal land points. The surface type, bare soil, is imposed at every land grid point of the climate model since land plants had yet to evolve. The ocean model bathymetry is a flat-bottom case (5000 m depth).

3.2 The Geochemical Model and the Coupling

The COMBINE model is an atmosphere (1 reservoir)-ocean (5 reservoirs) biogeochemical model originally coupled to an 1-D EBM [François and Walker, 1992] which is fully described in Goddéris and Joachimski [2003]. It includes the mathematical description of the geochemical cycles of carbon, phosphorus, alkalinity and oxygen. This model is designed to simulate long term (10⁵ to 10⁷ yrs) evolution of the global biogeochemical cycles.

Since a full coupling between COMBINE and CLIMBER-2 cannot be achieved due to excessive computation time, the numerical experiments were conducted in the following way: the climate model was run (with the environmental conditions described above) until equilibrium (5000 years) under an
atmospheric CO$_2$ concentration of 5000 ppm. Then starting from the previous equilibrium state each time, we conducted a suite of simulations (5000 years for each to reach the equilibrium) in which CO$_2$ level is held constant but is decreased in comparison with the previous one, until we reach the CO$_2$ level for which a global glaciation occurred. More than 40 simulations have been performed in this way in order to have, at least, a climate simulation for each 1°C of global cooling. We assume a linear behaviour in between each climatic simulations and thus, by doing an linear interpolation, we obtain the climatic variables of interest (temperature and runoff) for any pCO$_2$ required by the COMBINE geochemical model to calculate the consumption of CO$_2$ by continental weathering on a 7x18 grid. This consumption of CO$_2$ through silicate weathering ($F_{w,gra}$) is estimated using the following weathering law [Oliva et al., 2003]:

$$F_{w,gra} = k_{gra} \cdot \text{runoff} \cdot \text{area} \cdot \exp \left(-\frac{48200}{R} \left(\frac{1}{(T + 273.15)} - \frac{1}{288.15}\right)\right)$$

where runoff, area and $T$ are respectively the continental runoff, the continental area and the air temperature at the ground level for each of the continental grid elements. This law has been determined from a compilation study of the CO$_2$ ground level for each of the continental grid elements. This law run off, the continental area and the air temperature at the where

4.1 Overview of the Climatic Simulations

Each climatic steady-state simulation (with a fixed CO$_2$ value) is used to build a curve describing the evolution of the globally averaged, annual surface air temperature as a function of the atmospheric CO$_2$ level. In this way, two curves are plotted on figure 1a, one for the suite of NR runs and one for the suite of PAO runs. The behaviour of the CLIMBER-2 runs is consistent with other modelling studies as the ocean ice cover increases in response to the CO$_2$ decrease until the atmospheric CO$_2$ level reaches a threshold value that still yields an ice free tropical ocean. This threshold values are 57 and 78 ppm for the NR and the PAO experiments, respectively (see Fig. 2). Then, a further reduction of atmospheric pCO$_2$ drive the sea-ice line to the equator and the sudden transition from soft to hard snowball Earth takes a few hundreds years in both suite of experiments (Fig. 1b). These results show that the runaway ice-albedo feedback previously found in EBM’s [Budyko, 1969; Hyde et al., 2000] and in AGCMs [Jenkins and Smith, 1999; Pollard and Kasting, 2003; Poulsen et al., 2001] is also a prominent feature of the coupled ocean-atmosphere climate model CLIMBER-2 but due to the thermal inertia of the intermediate and deep ocean, the time scale of the transition is a few hundreds years rather than a few years.

The large mountain range of the PAO experiments induces a higher sensitivity to the radiative greenhouse gas forcing but the difference is rather small, the collapse occurring at 77 ppm for the PAO experiments instead of 56 ppm for the NR experiments. The topography is important for the initiation of glaciation essentially because of the atmospheric lapse rate promoting colder air temperatures and less summer melt (see below for a fully description of the topographic effect in these simulations). However, the applied topography is essentially located over the mid-to-high latitudes and thus does not strongly influence the way by which the tropical ocean freezes over.

The critical values required to maintain an open water solution calculated in this study are much lower than those found with AGCM studies, 840 ppm in Pollard and Kasting [2003],
700 ppm in Baum et al. [2002]. Although the different snow and sea-ice albedo values as well as the different ways of parameterising the ocean heat transport used in the climate models could explain the differences in sensitivity [Pierre-humbert, 2002], another factor implying a higher sensitivity of the AGCM simulations is the simple representation of the ocean in these models. The figure 3a shows the zonal sensible heat flux evolution over the ocean for the NR experiments at 500, 200, 100, 80 and 57 ppm. Once the sea ice margin migrates toward the subtropical latitudes (at 100 ppm, see Fig. 3a), the sensible heat flux displays a large increase in the front of the sea ice margin. The main source of sensible heat is the heat loss from the ocean to the atmosphere due to the colder air temperature surrounding the sea water close the sea ice. As a consequence, the heat flux to the ocean becomes more and more negative near the sea ice margin and the oceanic temperature shows a large decrease in the upper 500 m of the ocean with maximum changes at great depths occurring over the tropics between the NR experiments at 500 and 57 ppm (Fig. 3b). This tropical heat is stored in the upper 500 m and is continuously renewed by the large-scale advection, which transports the equatorial water toward the low latitudes. This heat transport slows down the migration of the sea ice line once the subtropical latitudes have been reached and explains the lower CO₂-sensitivity of the CLIMBER-2 model when

Figure 1. (A) Mean annual globally averaged surface air temperatures for various CO₂ levels from 2000 ppm to 56 ppm (77 ppm) for the ‘No Relief’ simulations, noted NR (‘Pan-African Orogen’ simulations, noted PAO). Note that each circle or square (NR and PAO runs, respectively) corresponds to a climate simulations in which the CO₂ level is held constant. (B) The mean annual surface air temperatures evolution over the first 1000 years of model integration for the largest atmospheric CO₂ levels resulting in a hard snowball Earth, 56 and 77 ppm for the NR and PAO experiments, respectively.

Figure 2. Geographical distribution of the mean annual sea ice at various CO₂ levels (A) for the NR experiments and (B) for the PAO experiments. The dashed grey line represents the location where the elevation is greater than 1000 meters. Note that the values, 57 and 78 ppm, are the lowest atmospheric CO₂ concentrations still yielding an open water solution (named a soft snowball Earth).
compared to the predictions of the AGCM coupled to a mixed layer representation of the oceans.

Although these results demonstrate that a full snowball Earth is a credible solution with a coupled ocean-atmosphere climate model, the mechanisms which might explain how such low atmospheric CO₂ levels might occur remain questionable and are now investigated.

4.2 No Relief Versus Pan-African Orogeny Experiments

We first run the geochemical model using the suite of NR climatic experiments and assuming a degassing rate of 6.810^{12} moles of carbon per year (required to balance the global consumption through the weathering of silicate lithologies under present day climatic conditions and continental configuration). A steady-state atmospheric pCO₂ of 917 ppm is achieved which is far away from the critical value of 56 ppm required to promote a 100% sea-ice cover. However, the resulting climate is rather cold given the reduction of 6% of the incoming solar radiation, being characterised by a mean global temperature of 10.11°C. There is no area on the mid-to-high latitudes continents below freezing in the austral summer as well as no area of permanent snow cover (Plate 1-a) although temperatures average −35°C in the interior regions during the austral winter (not shown).
Plate 1. Austral summer (mean of December, January and February) surface air temperatures (°C) in colour shading; contour lines are for the snow depth of 1 m, 5 m and 15 m surviving to the austral summer (note that 15 m is the maximum depth allowed to build up in the model). Surface winds are also shown for the latitudes 30°S–90°S, the size of the vector below the panel is for a wind of 8 m.s⁻¹. The dashed grey line represents the location where the elevation is greater than 1500 meters (and lower than 2000 meters) when accounting for the Pan-African Orogen. (A) Results for the NR experiments. (B) Results for the PAO experiments. (C) Results for the PAO experiments when accounting for the physical erosion (see text).
Adopting the reconstruction of the paleo-elevations of the East African orogen and of the Brazilian orogen used in Donnadieu et al. [2002], we re-run the climate/geochemical model. A higher equilibrated atmospheric pCO₂ of 1477 ppm is found and the mean global temperature reaches a value of 10.77°C. Nevertheless, areas of snow that survive to the austral summer appear in the southern parts of the mountain chains and in the coastal areas benefiting from the marine influence that keeps continental temperatures below freezing (Plate 1-b). In contrast, the interior region of the Laurentia craton close to the South pole experiences temperatures above 5°C as there is neither high relief nor the cooling effect of the marine influence that allows snow to build up. In fact, by adding the Pan-African belts, the atmospheric circulation at the southern mid-latitude is strongly modified. The mid-latitude westerlies blowing over the West Gondwana continent bring no more hot air masses over the oceanic surface around 0° of longitude as in the NoRelief simulation (Plate 1-a). As a consequence, a strong cooling occurs over this part of the ocean and is propagated over the north-western part of Laurentia via the westerlies which increase due to the larger zonal temperature gradient at these latitudes (compared Plate 1a & Plate 1b).

As a final remark, this strong cooling results in a drastic decrease of continental weathering rates in the north-western part of Laurentia and then induces a higher steady-state pCO₂ relative to the coupled climatic-geochemical NR experiments, since volcanic CO₂ degassing is assumed constant.

While we have observed the modification of climate due to the change in elevation of the land surface, the influence of the Pan-African uplift on the chemical weathering rates is more difficult to take into account as the orogen would probably increase the physical denudation of continental rocks through glacier abrasion and the development of steep slopes. A quantitative description of this mechanism is beyond the scope of this study. However, Millot et al, [2002] have established a strong positive correlation between the chemical weathering and the physical erosion for a dozen small granitic catchments, suggesting that CO₂ consumption through silicate weathering is a power law of the physical denudation rate. We thus apply a correction factor to the silicate weathering law following [Berner and Kothavla, 2001]:

\[ F_{\text{gra}}^{\text{w.i}} = k_{\text{gra}} \cdot \text{runoff} \cdot \text{area} \cdot \exp \left( -\frac{48200}{R} \cdot \left( \frac{1}{T + 273.15} - \frac{1}{288.15} \right) \right) \cdot \text{fmech}^{0.66} \]

The fmech factor is fixed to a constant value of 1 for pixels located outside of the Pan-African Orogen. For pixels located within the orogen, we assume a value of 3 for fmech. This latest number is based on the study of the modern analog of the Pan-African Orogen: the Himalayan uplift. Indeed Métivier et al. [1999] identified an increase of the sediment discharge originating from the Himalayan range by a factor of 3 over the course of the uplift, from the India-Asia collision 40 million years ago towards the present day. Finally, we assume that carbonate weathering is not influenced by physical erosion, because of the high solubility of carbonate minerals relative to silicate minerals.

This change in the continental weatherability drives down the atmospheric pCO₂ level to a steady-state value of 1037 ppm and results in a 2°C decrease of the global temperature that reaches 8.7°C. At mid-latitudes, the area of permanent snow cover migrates toward the south and the east of Laurentia due to the cool wind blowing over the continent (Plate 1-c). The global cooling also promotes a weak development of the perennial snow cover over the eastern part of the West Gondwana. However the high latitudes continental surfaces remain mainly influenced by a strong seasonality which prevents any accumulation of snow surviving to the austral summer except for the western part of the West Gondwana (Plate 1-c).

It appears that, under present day CO₂ degassing rate through volcanic activity, the geochemical forcing is sufficient to force the global climate into a colder state characterised by a regional extent of the permanent snow cover only if the Pan-African belts are included. Such cold conditions will facilitate the onset of a snowball event. However, the causal link existing between paleogeography, global climate and global carbon cycle may be hardly the culprit of the onset of a snowball glaciation during Marinoan times, since tropical temperatures remain well above freezing. In other terms, the atmospheric CO₂ level required to simulate a hard snowball Earth, 77 ppm, is still far away from being reached. However, nobody knows about the volcanic CO₂ degassing rate around 580 Ma. Therefore, we performed various coupled climatic-geochemical simulations by reducing the degassing rate, in order to check whether it would be possible to fall below the snowball CO₂ threshold of 77 ppm via an hypothetical decrease of this poorly constrained parameter. A CO₂ degassing rate reduced to 10% of its present day value results in a pCO₂ stabilizing around 170 ppm, corresponding to a global mean temperature of ~12°C (not shown), still far above the required 77 ppm. Despite the lack of constraints on the Neoproterozoic CO₂ degassing rate, such a drastic reduction is unlikely to have occurred: the degassing rate never reached so low values during the Phanerozoic [Engerbretson et al., 1992; Gaffin, 1987; Rowley, 2002], and we can furthermore expect a rough exponential decay of the CO₂ degassing flux since the formation of the Earth. This would imply higher CO₂ release from the mantle in the past, although this does not preclude periods of low degassing...
(related for instance to the presence of supercontinents, a configuration that does not fit the Marinoan times).

5 DISCUSSIONS AND CONCLUSIONS

A series of coupled ocean-atmosphere climate model simulations were carried out with a Late Neoproterozoic palaeogeography and solar luminosity reduced by 6% relative to today. Considerations of the long-term carbon cycle of the ocean-atmosphere system, together with the variations of the climate model predictions to changes in atmospheric pCO2, have led to the following conclusions.

First, by lowering sufficiently the atmospheric pCO2, 56 ppm and 77 ppm for the NR and the PAO experiments respectively, it is possible to initiate an ice-covered Earth even when accounting for the ocean dynamics as this is explicitly simulated in the CLIMBER-2 model. These results are not in conflict with the studies of Poulsen et al. [2001] who do not simulate an ice-covered Earth with the coupled ocean-atmosphere model FOAM, as the lowest values of atmospheric pCO2 they have used in their simulations is 140 ppm. In addition, the short duration of their experiments may do not allow them to account for a changes in ocean dynamics but only to the fast response of the surface layer in the tropical area.

Second, by using a geochemical-climate modelling approach to roughly estimate the atmospheric pCO2 levels, we show that adding mountain ranges to the 580 Ma continental configuration promotes a larger equilibrated pCO2 level (1477 ppm) than in the run with no relief (917 ppm). However, the NR run does not allow accumulation of perennial snow whereas the PAO run shows a strong modification of the atmospheric circulation due to the mountain ranges that results in a snow accumulation pattern compatible with a regional-scale glaciation. Furthermore, we have sought to test whether a large increase of the erosion flux, for instance due to physical erosion, may increase the silicate weathering sufficiently to drive the Earth to collapse into a hard snowball state. At best, we generate a regional-scale glaciation over the West Gondwana and the Laurentia cratons but the equilibrated pCO2 are still far removed from the critical values required to initiate an ice-covered Earth. These results could give additional support to the inferences of many authors that some of the glacial deposits originally attributed to a snowball-like glaciation could in fact be the consequence of a more localised glaciation. For instance, Barford et al. [2002] proposed that the Squantum and the Moelv glaciogenic deposits are more representative of a regionally extensive glaciation rather than of a hard snowball Earth. However, given the arbitrary assumption about the existence of a similar degassing rate at that time, and other modelling uncertainties concerning the weathering laws and the mountain ranges, these results have to be regarded as constituting only a crude sensitivity test of the sort of glaciations that may be triggered by accounting for the silicate weathering effect on the atmospheric CO2 content at the end of the Neoproterozoic. Particularly, non steady-state effects of orogenies on the global carbon cycle should be included in future versions of the geochemical module, such as the increasing burial of carbon within reduced sediments as a result of high sedimentation rate down to the uplifted area. Such processes have been suggested to constitute a non negligible carbon sink at the million year timescale in the particular case of the Himalayan uplift [France-Lanord and Derry, 1997].

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